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# Depth Migration of Seismovolcanic Tremor Sources Below the Klyuchevskoy Volcanic Group (Kamchatka) Determined From a Network-Based Analysis

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## Abstract

We present a method for automatic location of dominant sources of seismovolcanic tremor in 3D, based on the spatial coherence of the continuously recorded wavefield at a seismic network. We analyze 4.5 years of records from the seismic network at the Klyuchevskoy volcanic group in Kamchatka, Russia, when 4 volcanoes experienced tremor episodes. After enhancing the tremor signal with spectral whitening, we compute the daily cross-correlation functions related to the dominant tremor sources from the first eigenvector of the spectral covariance matrix, and infer their daily positions in 3D. We apply our technique to the tremors beneath Shiveluch, Klyuchevskoy, Tolbachik and Kizimen volcanoes, and observe the year-long pre-eruptive volcanic tremor beneath Klyuchevskoy from deep to shallow parts of the plumbing system. This observation of deep volcanic tremor sources demonstrates that the cross-correlation based method is a very powerful tool for volcano monitoring.

## Plain Language Summary

Volcanic tremors are the seismic signature of magmatic and hydrothermal fluids passing through volcanic conduits. Locating them in 3D is of real interest because it could allow us to monitor in more details movements of magma inside volcanic edifices and in some cases, to forecast eruptive episodes. The location of tremor in 3D nevertheless remains challenging because signals generated by tremors do not present any clear onset that could be used for picking arrival times and for determining source location. We design a method based on cross-correlations that recovers the differential travel times between receivers of a seismic network from the analysis of the statistically dominating waves in the wavefield. We present an application of the proposed method to volcanoes in Kamchatka, Russia, and show that we can track the pre-eruptive tremor episode in depth before the main eruption of the Klyushevskoy volcano.

## 1 Introduction

The long-period volcanic seismicity, composed of long-period events and volcanic tremors, is mainly thought to be related to processes induced by fluid movements within magmatic or hydrothermal volcanic systems (B. A. Chouet et al., 1996; Iverson et al., 2006). As it constitutes an important attribute of volcanic unrest, its detection and char-

acterization is a key aspect of volcano monitoring and eruption forecasting (McNutt, 1992; Sparks et al., 2012; B. A. Chouet & Matoza, 2013). Long-period earthquakes and volcanic tremors derive from similar processes but they differ in nature. Long-period earthquakes are discrete emergent signals of high-frequency onset followed by a low frequency harmonic waveform of typical frequency range 0.5 – 5 Hz (B. A. Chouet et al., 1996), that can sometimes become very frequent in episodes called "drumbeats". Volcanic tremors are highly irregular continuous signals that can last from a few minutes to several months, of which the spectral content can be harmonic, monochromatic, banded, spasmodic, etc. (Konstantinou & Schlindwein, 2003). However, both phenomena are believed to be generated by similar mechanisms such as magma moving through narrow cracks, fragmentation and pulsation of pressurized fluids within the volcano, or escape of pressurized steam and gases from fumaroles (B. A. Chouet & Matoza, 2013).

Long-period earthquakes are observed both in shallow parts of volcanoes (B. A. Chouet et al., 1996; Ciaramella et al., 2011; Firstov & Shakirova, 2014; Jolly et al., 2017; N. Shapiro et al., 2017; W. B. Frank et al., 2018; Woods et al., 2018, e.g.), as well as deep parts (Ukawa & Ohtake, 1987; Hasegawa et al., 1991; R. White et al., 1992; Pitt & Hill, 1994; R. A. White, 1996; N. Shapiro et al., 2017; W. B. Frank et al., 2018; Han et al., 2018; Hotovec-Ellis et al., 2018, e.g.), whereas most of reported volcanic tremor sources are superficial (< 5 km). Indeed, except the deep volcanic tremors (30–60 km) observed at Kilauea volcano between 1962 and 1983 (Aki & Koyanagi, 1981; Shaw & Chouet, 1991), the majority of volcanic tremor studies deal with superficial sources. Thus, shallow volcanic tremor sources are reported on eruptive craters and conduits (Goldstein & Chouet, 1994; Métaxian et al., 2002; Battaglia et al., 2005; Di Grazia et al., 2006; Almendros et al., 2014; Ogiso et al., 2015; Ichimura et al., 2018; Moschella et al., 2018, e.g.), building cones (Haney, 2010), gas slug bursts at surface vents (B. Chouet et al., 1997; Ripepe et al., 2001; Fee et al., 2010; Patrick et al., 2011; Yukutake et al., 2017; Barriere et al., 2017; Eibl, Bean, Jónsdóttir, et al., 2017, e.g.), lava lakes (Métaxian et al., 1997; Barriere et al., 2017; Donaldson et al., 2017, e.g.), lava flows (Caudron et al., 2015), shallow dikes (Eibl, Bean, Vogfjörd, et al., 2017; Eibl, Bean, Jónsdóttir, et al., 2017; Woods et al., 2018, e.g.), magma-water aquifer interfaces (Almendros et al., 1997), etc.

Methods currently used for volcanic tremor source location can be divided into following categories: small aperture array methods (Goldstein & Chouet, 1994; Almendros et al., 1997; B. Chouet et al., 1997; Métaxian et al., 2002; Almendros et al., 2007; Haney,

2010; Almendros et al., 2014; Eibl, Bean, Vogfjörd, et al., 2017; Eibl, Bean, Jónsdóttir, et al., 2017, e.g.), network-based amplitude decay methods (Aki & Koyanagi, 1981; Ripepe et al., 2001; Battaglia et al., 2005; Di Grazia et al., 2006; Taisne et al., 2011; Ogiso et al., 2015; Ichimura et al., 2018; Moschella et al., 2018, e.g.), and network-based inter-stations cross-correlations methods (Ballmer et al., 2013; Droznin et al., 2015; Li et al., 2016, 2017; Donaldson et al., 2017; Barriere et al., 2017; Yukutake et al., 2017; Woods et al., 2018, e.g.). It is interesting to mention that network-migration-based methods have been also used to locate tectonic tremor and low-frequency earthquakes (Kao & Shan, 2004; W. Frank & Shapiro, 2014, e.g.). Dense small aperture arrays are not always available because of their cost and their temporal nature, but large scale seismic networks are nowadays permanently deployed in many places. Because of assumptions about geometrical spreading and attenuation, network-based amplitude decay methods are adapted to locate volcanic tremor sources relatively close to the used network, i.e. sources that are not so far or not so deep. On the other hand, inter-stations cross-correlations methods can observe tremors at larger scales (Ballmer et al., 2013) and might be able to locate deeper sources. Previous volcanic tremor location studies based on cross-correlations have located shallow tremor sources at Kilauea (Ballmer et al., 2013; Donaldson et al., 2017), Katla (Li et al., 2016, 2017), Nyiragongo and Nyamulagira (Barriere et al., 2017), Hakone (Yukutake et al., 2017) and Bardarbunga (Woods et al., 2018) volcanoes.

To our knowledge, no deep volcanic tremor source has been located with this method so far. This is certainly due to the fact that a surface wave propagation is often assumed for the analysis, inherently implying the source to be located in the shallow part. The automatic cross-correlations based location method proposed in our article is a 3D approach based on the assumption of dominant S-wave emission from tremor sources and associated S-wave structures, enabling to look for both shallow and deep tremors. By using the daily first eigenvector of the network covariance matrix, we are able to track the temporal evolution on a daily time scale of the 3D location of the daily dominating volcanic tremor sources.

We apply the developed method to data continuously recorded by the permanent monitoring network of the Klyuchevskoy group of volcanoes (Kamchatka, Russia, see Fig. 1) during a four and a half years-long time period between January 2009 and June 2013. During this period of time, Klyuchevskoy, Tolbachik, Kizimen and Shiveluch volcanoes experienced episodes of volcanic tremor (Droznin et al., 2015) and abundant amount of

118 long-period earthquakes (N. Shapiro et al., 2017). The two most important tremors led  
119 to a summit eruption at the end of 2010 at Klyuchevskoy (Senyukov, 2013, e.g.) and a  
120 fissure eruption beginning at the end of 2012 at Tolbachik (Gordeev et al., 2013, e.g.).  
121 Short explosive eruptions also occurred at Bezymianny. The seismic monitoring network  
122 and volcanic tremors acting in the studied volcanic group are described in Sec. 2. Sec 3  
123 details the method developed to automatically locate tremor sources in 3D. The tem-  
124 poral evolution of those 3D locations is described and discussed in Sec. 3.3 and 4.

## 125 **2 Volcanic Tremors at the Klyuchevskoy Volcanic Group**

126 The Klyuchevskoy volcanic group (KVG) is one of the most active subduction-zone  
127 volcanic group in the world. It consists of 13 strato-volcanoes located in an approxima-  
128 tively 70 km diameter zone on the Russian Kamchatka peninsula (N. M. Shapiro et al.,  
129 2017). Three volcanoes of the KVG (Klyuchevskoy, Bezymianny and Tolbachik) were  
130 active during recent decades (N. M. Shapiro et al., 2017; Gordeev et al., 1989; Ozerov  
131 et al., 2007; Ivanov, 2008; Senyukov et al., 2009; Senyukov, 2013, e.g.) . In addition, two  
132 other very active volcanoes, Shiveluch and Kizimen, are respectively located North and  
133 of KVG (Fig. 1).

### 134 **2.1 Single-station based monitoring of volcanic tremors**

135 The Kamchatka Branch of the Geophysical Survey (KBGS) of the Russian Academy  
136 of Sciences operates the permanent seismic network monitoring the KVG and surround-  
137 ing volcanoes (Gordeev et al., 2006; Chebrov et al., 2013). In this paper, we use four and  
138 a half years of seismic data continuously recorded between January 2009 and June 2013  
139 by this permanent network composed of 18 stations in 2009 and 19 stations from 2010  
140 onwards. The stations and the five monitored volcanoes are shown on Fig. 1. The seis-  
141 mic stations have three-components, each one being equipped by a CM-3 short period  
142 seismometer with a corner frequency of around 0.8 Hz. Continuous records are sampled  
143 at 128 Hz and only the vertical component is analysed in this article, studied tremors be-  
144 ing more energetic in the vertical direction (but horizontal components could also be used  
145 as mentioned in Sec. 2.2.1).

146 As many volcanological observatories do all over the world, the KBGS monitors  
147 volcanic tremors using a simple single-station approach. For example, stations LGN and

148 KMN are used to monitor Klyuchevskoy and Tolbachik tremors respectively (Fig. 1). This  
 149 simple approach suffers from limitations of robustness if the reference station breaks down,  
 150 precision if electronic noise is confused with tremor, and impossibility to spatially locate  
 151 the tremor.

## 152 **2.2 Network-based monitoring of volcanic tremors**

153 To overcome the limitations of single-station based approaches, we propose to use  
 154 a network-based strategy that exploits the coherence of signals simultaneously recorded  
 155 by several receivers. It is based on the network covariance matrix that can be used to  
 156 extract coherent signals propagating across the network by reducing local noise (Seydoux,  
 157 Shapiro, de Rosny, et al., 2016). A detailed explanation of the covariance matrix esti-  
 158 mation from real data is presented in Sec. 2.2.1.

159 Being interested on long-duration tremors and not on short-transient events, all in-  
 160 ferences are obtained from day-long network covariance matrices. Once detected through  
 161 the analysis of the daily network covariance matrix eigenvalues (please refer to Sec. 2.2.2  
 162 for more information), different volcanic tremors can be clustered and located by using  
 163 the daily network covariance matrix first eigenvector (Sec. 2.2.3).

### 164 ***2.2.1 Estimation of the daily network covariance matrix***

165 We here narrow the focus on the vertical-component records in our analysis, though  
 166 an extension of this strategy to 3-component data is possible (Wagner & Owens, 1995).  
 167 We form the spectral network data vector  $\mathbf{u}(f, t)$  from the records of the  $N$  seismic sta-  
 168 tions as

$$169 \quad \mathbf{u}(f, t) = [u_1(f, t), u_2(f, t), \dots, u_N(f, t)]^T, \quad (1)$$

170 where  $u_i(f, t)$  is the Fourier transform of the signal recorded at the station  $i = 1 \dots N$   
 171 calculated over a time window of duration  $\delta t$  and starting at  $t$ .

172 The general workflow of covariance matrix estimation is summarized with the next  
 173 steps:

- 174 1. Seismic data are downsampled from 128 Hz down to 25.6 Hz after bandpass-filtering  
 175 between 0.1 and 10 Hz in order to allow for faster computation. We consider the

- 176 raw data without correcting for the instrument response, because all the sensors  
177 are identical.
- 178 2. We apply spectral whitening to the seismic records (independently for every sta-  
179 tion) in order to enhance the coherence of the tremors signals (Bensen et al., 2007).  
180 Spectral whitening is often combined with temporal normalization in order to at-  
181 tenuate the amplitude of strong impulsive sources (earthquakes). But applied alone,  
182 as done here, the spectral whitening has shown to significantly improve the de-  
183 tection of volcanic tremors at Kamchatka (Soubestre et al., 2018). Indeed, it di-  
184 minishes the influence of strong earthquakes and local noise and preserves the am-  
185 plitude modulation of tremor signals, that is a useful information when tremors  
186 are not fully stationary in time and rather represent sequences of many impulsive  
187 long-period events.
- 188 3. We divide the seismic data into time windows of duration  $\delta t$  from which the Fourier  
189 transform is calculated in order to obtain the spectral data vector (Eq. (1)). We  
190 also avoid artefacts caused by sharp windows by applying a tapering Hann win-  
191 dow; we thus overlap the signal segments with a factor of 50% in order not to loose  
192 any information located in consecutive windows.
- 193 4. We calculate the covariance matrix from sets of  $M$  consecutive overlapping seg-  
194 ments of data, such as

$$195 \quad \mathbf{C}(f, t) = \langle \mathbf{u}(f, t) \mathbf{u}^\dagger(f, t) \rangle_{\Delta t} = \frac{1}{M} \sum_{m=0}^{M-1} \mathbf{u}(f, t + m\delta t/2) \mathbf{u}^\dagger(f, t + m\delta t/2). \quad (2)$$

- 196 5. The covariance matrix  $\mathbf{C}(f, t)$  is then obtained for each time  $t$  and incorporates  
197 information about the data collected in the interval  $[t, t + \Delta t]$  where  $\Delta t = (1 +$   
198  $M)\delta t/2$ . We decide to calculate the covariance matrix with a sliding step of  $\Delta t/4$   
199 in order to have an appreciable final temporal resolution in our analysis.
- 200 6. The daily covariance matrix is finally obtained by averaging consecutive covari-  
201 ance matrices obtained within one day of data. Note that this windowing and av-  
202 eraging is necessary because a matrix built from day-long spectra (without win-  
203 dowing) would be generated from a single data vector. Inherently, such a matrix  
204 being of rank one, only a single eigenvector could be obtained. In order to observe  
205 the splitting of different sources onto different eigenvectors, we must average the  
206 wavefield observations onto smaller windows, allowing to increase the rank of the  
207 covariance matrix and to better separate sources.

208 This daily network covariance matrix being inherently Hermitian and positive semi-  
 209 definite (Seydoux, Shapiro, de Rosny, et al., 2016), it can be decomposed on the basis  
 210 of its complex eigenvectors  $\mathbf{v}^{(i)}$  associated with real positive eigenvalues  $\lambda_i$ :

$$211 \quad \mathbf{C}(f, t) = \sum_{i=1}^N \lambda_i(f, t) \mathbf{v}^{(i)}(f, t) \mathbf{v}^{(i)\dagger}(f, t) \quad (3)$$

### 212 *2.2.2 Detection of tremors from the covariance matrix eigenvalues*

213 According to Seydoux, Shapiro, de Rosny, et al. (2016), the number of non-zero eigen-  
 214 values is related to the number of statistically independent signals composing the wave-  
 215 field and then of independent seismic sources. Real seismic data is composed by a large  
 216 amount of seismic sources ranging from earthquakes, tremors, and more complex sources  
 217 usually designated as noise sources (hum, oceanic microseisms, wind, anthropic activ-  
 218 ity, scatterers acting like secondary sources, self-sensor electronic noise...). This large num-  
 219 ber of sources are not absolutely independent and may be located onto similar eigenvec-  
 220 tors. Also, the estimation of the covariance matrix from a finite number of windows may  
 221 also reduce our ability to capture the independent seismic signals. For these reasons, de-  
 222 riving the exact number of independent sources is a challenge (Wax & Kailath, 1985, e.g.,)  
 223 that is out of scope in the present study. Nevertheless, an estimate of the “level of co-  
 224 herence” of the seismic wavefield can be obtained by calculating the width  $\sigma(f, t)$  of the  
 225 covariance matrix eigenspectrum (the covariance matrix *spectral width*), a proxy for the  
 226 presence of coherent signal within the seismic wavefield:

$$227 \quad \sigma(f, t) = \frac{\sum_{i=1}^N (i-1) \lambda_i(f, t)}{\sum_{i=1}^N \lambda_i(f, t)} \quad (4)$$

228 This spectral width can be seen as a proxy for the number of independent seismic sources.  
 229 Thus, the spectral width corresponding to ambient seismic noise produced by distributed  
 230 noise sources is high, whereas the spectral width of a signal spatially coherent at the net-  
 231 work scale generated by a single localized source is low.

232 The spectral width  $\sigma(f, t)$  computed from daily covariance matrices during the 4.5  
 233 years-long studied time period is analyzed by Soubestre et al. (2018). This result clearly  
 234 shows three main long-duration episodes when the spectral width is reduced: (1) in the  
 235 beginning of 2009, (2) from the end of 2009 to the end of 2010, and (3) from the end of  
 236 2012 to mid 2013. From *a priori* knowledge (Droznin et al., 2015), we know that the two  
 237 first episodes correspond to periods of strong non-eruptive-tremors at Klyuchevskoy and  
 238 the last one to a strong eruptive tremor at Tolbachik. This approach based on eigenval-

239 ues is therefore efficient to detect volcanic tremor, but it is unable to tell whether those  
 240 detected tremor sources are from the same volcano or from different ones (here we know  
 241 it from *a priori* knowledge). To overcome the limitation of this approach to distinguish  
 242 different sources of volcanic tremor, Sec. 2.2.3 focuses on the eigenvectors of the daily  
 243 network covariance matrix.

### 244 ***2.2.3 Clustering and location of tremors from the covariance matrix first*** 245 ***eigenvector***

246 Instead of using directly the daily network covariance matrix to cluster or locate  
 247 tremor sources, it has been shown in Soubestre et al. (2018) that better results are ob-  
 248 tained by using the covariance matrix first eigenvector. The idea is derived from prin-  
 249 cipal component analysis, where the first component (eigenvector) is known to represent  
 250 the most coherent part of the wavefield (Wagner & Owens, 1995; Seydoux, Shapiro, Rosny,  
 251 & Landès, 2016). Therefore, for days with tremor activity, the first eigenvector charac-  
 252 terizes the dominant tremor source, and acts as a denoising operator because the higher-  
 253 order eigenvectors are orthogonal to the first one, and most likely span information re-  
 254 lated to the background noise. It has to be noticed that such a filtering by the first eigen-  
 255 vector theoretically avoids the possibility of identifying simultaneously active tremor sources  
 256 from different volcanoes. Being mainly interested by the daily dominating tremor source  
 257 in the studied region, this is not a problem for our purpose. But further work should be  
 258 realized to analyze in detail periods when more than one volcano are emitting tremor  
 259 simultaneously, what is out of scope of the present article. This distinction of simulta-  
 260 neously acting tremors might require to include higher-order eigenvectors of the covari-  
 261 ance matrix in the analysis. Another option could be to consider horizontal components,  
 262 in order to bring polarization constraint in the analysis (Vidale, 1986).

263 The frequency-dependent first eigenvector of the daily covariance matrix  $\mathbf{V}_1(day, f)$   
 264 can be analysed to cluster and/or locate the daily dominant tremor source. Using cor-  
 265 relation coefficient as a measure of their similarity, Soubestre et al. (2018) developed a  
 266 clustering algorithm to analyze this collection of daily first eigenvectors. The clustering  
 267 process identifies seven clusters associated with seismo-volcanic activity during the time  
 268 period considered in the present article. Each cluster's central day eigenvector (highest  
 269 correlation coefficient) corresponds to the most characteristic first eigenvector of the clus-  
 270 ter. Those characteristic eigenvectors can then be used as templates for template match-

271 ing detection of tremor sources, comparing them with the daily first eigenvectors from  
 272 the continuous records (Droznin et al., 2015) or with new available data. This cluster-  
 273 ing approach is interesting to classify different sources of volcanic tremor, but it does not  
 274 associate the obtained clusters to volcanoes of the studied zone. This location of tremors  
 275 can be done using the moveout information of the daily first eigenvector, as detailed in  
 276 the next section.

### 277 **3 Automatic 3D Location of Volcanic Tremor Sources**

278 Methods of volcanic tremor sources location from network cross-correlations have  
 279 already been proposed in previous studies (Ballmer et al., 2013; Droznin et al., 2015; Li  
 280 et al., 2016, 2017; Donaldson et al., 2017; Barriere et al., 2017; Yukutake et al., 2017; Woods  
 281 et al., 2018, e.g.), but with an *a priori* on the type of wave emerging in cross-correlations.  
 282 Because they study shallow tremor sources, the authors assume that cross-correlations  
 283 are dominated by surface waves. Conversely to those methods considering a surface wave  
 284 propagation from the tremor source, the cross-correlations based location method de-  
 285 veloped in our article assumes an S-wave propagation and the knowledge of the associ-  
 286 ated S-wave structure, allowing to locate shallow as well as deep tremor sources. Another  
 287 main difference between the method proposed here and existing ones, is that the con-  
 288 sidered cross-correlations are not directly those of the inter-stations cross-correlations  
 289 matrix, but those derived from the first eigenvector of the network covariance matrix.  
 290 This makes the cross-correlations more representative of the dominant tremor, as proved  
 291 in the supporting information S1 where the location of a synthetic tremor (made of mul-  
 292 tiple long-period events merging) is more precisely recovered using the filtering of the  
 293 network covariance matrix by its first eigenvector than the full matrix itself.

#### 294 **3.1 Method: 3D location from cross-correlations envelopes**

295 Volcanic tremor sources can be located using the moveout information contained  
 296 in the daily network covariance matrix first eigenvector, that contains an information  
 297 about the location and the mechanism of the dominating tremor source. We extract the  
 298 filtered covariance matrix  $\tilde{\mathbf{C}}(f, t)$  from the complex outer product of the first eigenvec-  
 299 tor  $\mathbf{v}^{(1)}(f, t)$  with its Hermitian transpose

$$300 \quad \tilde{\mathbf{C}}(f, t) = \mathbf{v}^{(1)}(f, t)\mathbf{v}^{(1)\dagger}(f, t). \quad (5)$$

301 This operation can be seen as a low-rank denoising of the covariance matrix, a strategy  
 302 widely used in image processing to remove the noise spanned by higher-order eigenvec-  
 303 tors (Orchard et al., 2008). The inverse Fourier transform of this matrix retrieves the  
 304 time-domain filtered cross-correlations:

$$305 \quad \tilde{\mathbf{R}}(\tau, t) = \mathcal{F}^{-1} \tilde{\mathbf{C}}(f, t), \quad (6)$$

306 where  $\mathcal{F}^{-1}$  is the inverse Fourier transform operated to the frequency dimension, and  
 307  $\tau$  is the cross-correlation time lag.

308 To locate the position of the dominant tremor source we use a grid search on a 3D  
 309 space grid. We compute covariance matrices (5) at frequencies between 0.5 Hz and 2.0 Hz  
 310 (spectral band where tremors are more energetic in Kamchatka according to Droznin et  
 311 al. (2015) and Soubestre et al. (2018)). The resulting time-domain cross-correlations are  
 312 naturally filtered in the same range. We then compute the smooth envelope of the cross-  
 313 correlations  $\mathbf{E}(\tau, t)$  by computing the absolute value of the analytic signal derived from  
 314 the Hilbert transform  $h$ , and performing a convolution with a Gaussian filter  $g(t)$  with  
 315 a 100 seconds width:

$$316 \quad \mathbf{E}(\tau, t) = g(t) \otimes |\tilde{\mathbf{R}}(\tau, t) + ih(\tilde{\mathbf{R}})(\tau, t)| \quad (7)$$

317 where,  $g(t) = \frac{1}{\beta\sqrt{2\pi}} \exp\left(-\frac{1}{2} \frac{t^2}{\beta^2}\right)$  is a Gaussian window with typical width  $\beta = 10$  seconds  
 318 (larger values have been tested with no significant difference and the  $\beta = 10$  s value was  
 319 found to be the best lower bound). The smoothing operation allows to reduce the alter-  
 320 ation of the location quality from scattering due to local heterogeneities and from the  
 321 imprecision of the location method itself, particularly the imprecision of the velocity model  
 322 described hereafter.

323 Then, for every point  $\mathbf{r} = x\mathbf{e}_x + y\mathbf{e}_y + z\mathbf{e}_z$  of the 3D grid, each cross-correlation  
 324 envelope  $E_{ij}(\tau)$  is shifted by the time difference  $d\tau_{ij}(\mathbf{r})$  between travel times needed for  
 325 an S-wave to travel from the tested point to the two stations  $i$  and  $j$ :

$$326 \quad d\tau_{ij}(\mathbf{r}) = \tau_i(\mathbf{r}) - \tau_j(\mathbf{r}) \quad (8)$$

327 Travel times  $\tau_i(\mathbf{r})$  and  $\tau_j(\mathbf{r})$  are calculated based on the 1D velocity model for S-waves  
 328 (dominating in volcanic tremors) accounting for velocity variation with depth derived  
 329 by Droznina et al. (2017), described in supporting information S2. The value at zero lag-  
 330 time of shifted cross-correlations envelopes is finally stacked for all stations pairs to ob-  
 331 tain the network response function and this computation is repeated for every day start-

ing at time  $t$ :

$$b(\mathbf{r}, t) = \sum_{i=1}^N \sum_{j>i}^N E_{ij}(-d\tau_{ij}(\mathbf{r}), t) \quad (9)$$

Note that cross-correlations envelopes being not relatively normalized, the weight in the stack of the network response of each individual cross-correlation envelope is directly related to its quality (spiky shape).

The grid search is performed on a grid covering the whole studied zone down to 50 km depth, with a 1 km  $\times$  1 km  $\times$  1 km resolution. In order to compare relative levels of tremor between different days, each daily network response function  $b(\mathbf{r}, t)$  is normalized first by its integral on the total grid volume  $V$ :

$$\ell(\mathbf{r}, t) = \frac{b(\mathbf{r}, t)}{\int_V b(\mathbf{r}, t) d\mathbf{r}} \quad (10)$$

Then, those spatially normalized response functions are temporally re-normalized by the maximum of the whole time period:

$$\tilde{\ell}(\mathbf{r}, t) = \frac{\ell(\mathbf{r}, t)}{\max_t \ell(\mathbf{r}, t)} \quad (11)$$

The integral-normalized network response (10) summing to unity on the total grid volume, it can be referred to as a presence likelihood of the tremor source. Similarly, the re-normalized response (11) represents this presence likelihood relatively to the maximum likelihood of the considered period. The most likely daily source position corresponds to the maximum of this function in space.

### 3.2 Results: 3D location of volcanic tremors

Fig. 2 shows examples of renormalized network response function (11) for selected days. Most probable daily tremor source location is marked by a black star. This figure clearly shows that during the studied time period, volcanic tremors are originating from different volcanoes or from different activity phases of a particular volcano. Thus, Fig. 2a shows a day of quite deep tremor activity beneath Shiveluch volcano. Fig. 2b and Fig. 2c respectively correspond to a deep and a shallow tremor activity beneath Klyuchevskoy. Fig. 2d and Fig. 2f respectively show days of shallow tremor activity beneath Kizimen and Tolbachik. Finally, a response function corresponding to a day during the calm period in 2012 without tremor activity, but maybe dominated by a regional earthquake, appears in Fig. 2e for comparison. For more information, supporting information S3 shows cross-correlations envelopes corresponding to Fig. 2c and Fig. 2e.

362 If some days correspond to shallow tremor sources (Figs. 2c, 2d and 2f), other days  
 363 such as 2009-088 of Fig. 2a and 2009-276 of Fig. 2b are showing a deep tremor source.  
 364 This result is quite interesting, because it is the first time that a deep volcanic tremor  
 365 is located by a cross-correlation based method. It means that during those days, the lo-  
 366 cated volcanic tremor has a source associated with body waves. In the context of the KVG,  
 367 those body waves are emitted by volcanic tremor sources located in the consolidated bedrock  
 368 under unconsolidated sedimentary or volcanic deposits. Note that such an emergence of  
 369 body waves in cross-correlations has already been observed in other different contexts  
 370 and scales, such as body waves propagating through the active magmatic system of the  
 371 Piton de la Fournaise volcano (Nakata et al., 2016), body waves reflected from the Moho  
 372 (Poli, Pedersen, & Campillo, 2012) or body waves from the mantle transition zone (Poli,  
 373 Campillo, et al., 2012).

### 374 **3.3 Results: Time evolution of volcanic tremor sources 3D location**

375 The issue of the vertical migration of volcanic tremor sources with time is addressed  
 376 here. The tremor presence likelihood (11) is daily analyzed beneath each studied vol-  
 377 cano (Shiveluch, Klyuchevskoy, Tolbachik and Kizimen from North to South) to check  
 378 for the source depth. To do so, a circular zone of a given radius is defined around each  
 379 volcano (the zone is centered on volcanoes summit as appearing on Fig. 1) and the max-  
 380 imum value of the likelihood (11) over those circular zones is selected at every depth and  
 381 every day. The expected value  $\bar{z}$  of the likelihood for each zone and each time is calcu-  
 382 lated with  $\bar{z}(t) = \int \max_{x,y} \tilde{\ell}(\sqrt{x^2 + y^2} < r, z, t) z dz$  and represented as a black dot in  
 383 order to better visualize the expected position of the tremor sources. Fig. 3 shows the  
 384 result of such an analysis for 25 km radius zones, as shown in Fig. 1. Note that the like-  
 385 lihood has been clipped below a value of 0.7 in order to narrow the focus on the highly-  
 386 coherent time segments.

387 This time-depth analysis clearly enhances the same time periods of tremor activ-  
 388 ity as the detection of tremors by Droznin et al. (2015) and Soubestre et al. (2018), but  
 389 with the benefit that here each tremor is clearly associated to a specific volcano. Dis-  
 390 regarding the depth information and looking only at the time axis of Fig. 3, this result  
 391 showing when each volcano is under tremor is interesting from a monitoring point of view  
 392 and this approach could be usefully run in real-time in a volcanological observatory. Thus,  
 393 volcanic tremor activity is automatically located at Shiveluch volcano in March 2009,

394 at Klyuchevskoy volcano at the beginning of 2009 and also from September 2009 to the  
395 end of 2010, at Kizimen during two episodes of one month in June and September 2011,  
396 and at Tolbachik volcano from the end of November 2012 onwards. We also see that dur-  
397 ing periods of volcanic rest (e.g. March 2011) the likelihood presence is the highest at  
398 great depth, and is most likely related to strong earthquakes that resisted our prepro-  
399 cessing; this is mostly due to the presence of dominating body waves in the seismic wave-  
400 field at global scales usually observed after big events (Boué et al., 2014; Sens-Schönfelder  
401 et al., 2015). For example in this period, the M9.1 Tohoku earthquake (Ozawa et al., 2011)  
402 is clearly visible on our results for several days (Fig. 3, beneath the Tolbachik inset).

403 Moreover, the depth of the daily dominating tremor source is made better visible  
404 by Fig. 3. Looking at the spatial distribution of volcanoes and seismic stations in the  
405 map of Fig. 1, one understands that the source depth is not well constrained at volca-  
406 noes Shiveluch and Kizimen that are located at the border of the studied zone and count  
407 with a few stations. Likewise, Tolbachik volcano is more central in the map but it counts  
408 only one station close to it, so that the depth estimation of tremor sources beneath Tol-  
409 bachik cannot be very precise. Fig. 3 shows that Tolbachik’s tremor is an eruptive tremor,  
410 beginning the 27th of November 2012 one day after the official beginning of the fissure  
411 eruption that started the 26th of November. Looking at Fig. 2f, the tremor source seems  
412 to be superficial. This result is consistent with the analysis from Caudron et al. (2015)  
413 who relate this eruptive tremor to lava flows. This is not surprising for this tremor to  
414 be superficial given that the seismic preparatory phase of Tolbachik’s 2012-2013 erup-  
415 tion was also shallow. Indeed, the eruption was preceded by 4-5 months of shallow (depth  
416  $< 5$  km) and low-energy volcano-tectonic seismicity (Kugaenko et al., 2015), as well as  
417 shallow long-period events clustered in time due to multiple cracks and channels reac-  
418 tivation (W. B. Frank et al., 2018). The eruption is supposed to have been fed by magma  
419 ascending from a shallow crustal magma storage (Fedotov et al., 2010, 2011).

420 On the other hand, the depth of tremor sources is better constrained for Klyuchevskoy  
421 volcano that is central in the map and count multiple stations around it. Moreover, the  
422 plumbing system of Klyuchevskoy is known to be more complex and to contain deeper  
423 levels of magma storage, as evidenced by different types of studies. Thus, Koulakov et  
424 al. (2011) imaged a crustal storage beneath Klyucevskoy at intermediate depth of 8-13 km  
425 and a deeper one in the mantle below 25 km depth. This latter deep storage was also iden-  
426 tified by analysis of deep long-period earthquakes (N. Shapiro et al., 2017). In Fig. 3,

427 the 2009-2010 pre-eruptive tremor of Klyuchevskoy volcano seems to affect shallow (depth  
428  $< 5$  km), medium ( $5$  km  $<$  depth  $< 15$  km) as well as deeper depths of about 20-25 km  
429 at the beginning of the time period. The tremor shows different phases from Septem-  
430 ber 2009 to December 2010. First, from September to December 2009, the expected po-  
431 sition of the tremor source (black dots in Fig. 3) progressively migrates from deep depths  
432 of about 25 km to medium depths of 10 km. Then, the tremor activity seems to stabi-  
433 lize around 10 km from January to June 2010, when it starts to focus to shallower depths  
434 of a few kilometres. At the beginning of July 2010 the expected position of the tremor  
435 source starts to move deeper again down to 15 km, where it stabilizes up to the end of  
436 October 2010. This change could be interpreted as a possible reinjection of magma from  
437 a deep reservoir. And eventually, from November 2010 onwards the dominant tremor source  
438 progressively migrates to the surface, a few months before the beginning of the summit  
439 eruption that started in December 2010. This evidence of the sequence of dominant tremor  
440 source depth decrease with time getting closer to the eruption is an interesting result con-  
441 sistent with the dynamics of magmatic eruption processes.

#### 442 **4 Conclusion**

443 In this paper, we developed a cross-correlations based method to automatically lo-  
444 cate volcanic tremor sources in 3D. The method consists in a first preprocessing step to  
445 normalize the data and calculate the daily network covariance matrix, a second step of  
446 filtering this matrix by its first eigenvector to retrieve the daily dominant tremor source,  
447 and a last step to locate this source from the moveout information of the first eigenvec-  
448 tor. An important aspect of the developed method is that once the S-wave velocity model  
449 and all parameters selected, the data analysis and the following detection and location  
450 of tremor sources are fully automatic without need of additional *a priori* information.  
451 A main difference between the developed cross-correlations based location method of vol-  
452 canic tremor and other existing ones, is that instead of assuming a surface wave type prop-  
453 agation and a shallow origin of the tremor, we estimate the tremor source depth and we  
454 are able to locate shallow as well as deep tremors. Another main difference is that the  
455 considered cross-correlations are not directly those of the inter-stations cross-correlations  
456 matrix, but those derived from the first eigenvector of the network covariance matrix that  
457 characterizes the dominant component of the recorded wavefield.

458 The daily application of the developed method onto four and a half years of data  
459 (from January 2009 to June 2013) from the Klyuchevskoy group of volcanoes allowed  
460 to daily locate the dominant volcanic tremor source, what is interesting from the mon-  
461 itoring point of view of a volcanological observatory. It also enabled to track the tem-  
462 poral evolution of the 3D location of the pre-eruptive volcanic tremor beneath Klyuchevskoy  
463 from deep to shallow parts of the plumbing system, from September 2009 to December  
464 2010 when the summit eruption started. This latter result is interesting, first because  
465 to our knowledge it is the first time a deep volcanic tremor source is located with a cross-  
466 correlation based method. Second, because the evolution of the tremor source depth dur-  
467 ing one year and half including pre and co-eruptive activity is quite nicely recovered. Ac-  
468 cording to the analysis, initially multiple tremor sources are acting simultaneously from  
469 the surface to depths down to 20-25 km, and then the dominating tremor source progres-  
470 sively migrates to the surface. This surface migration pattern is observed twice during  
471 the 2009-2010 tremor of Klyuchevskoy, first from September 2009 to the end of June 2010  
472 and second from mid July to December 2010 after a possible re-injection in July-August.  
473 This behavior of volcanic tremor sources affecting a wide depth range and then migrat-  
474 ing to the surface can be explained by the model proposed by Stasiuk et al. (1993) of  
475 progressive magma draining from the bottom of vertical pathways during eruptions.

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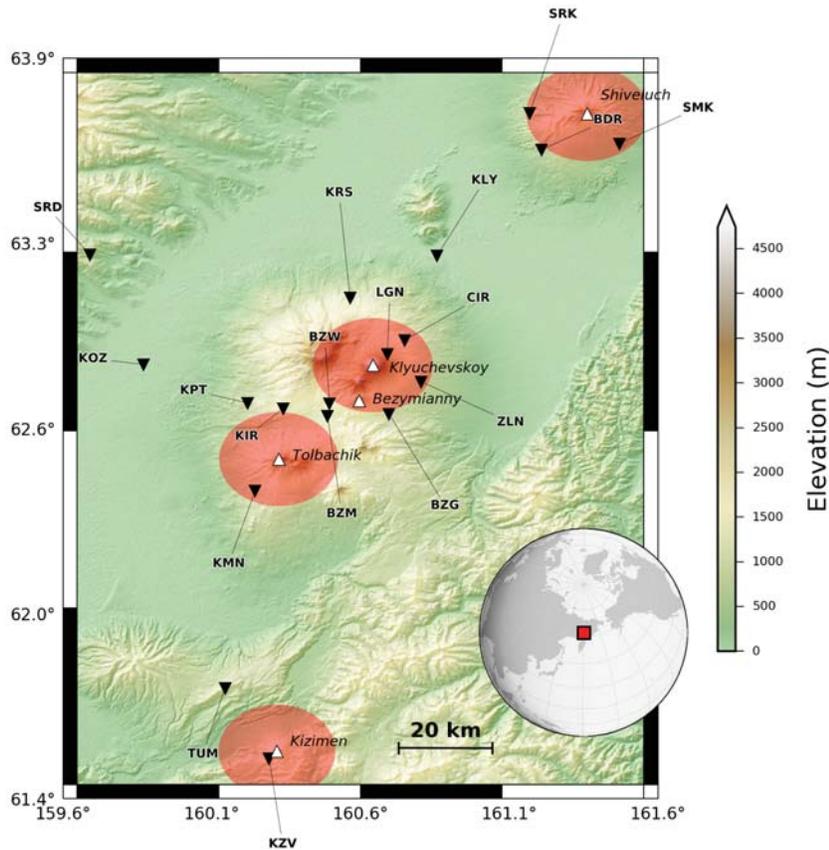
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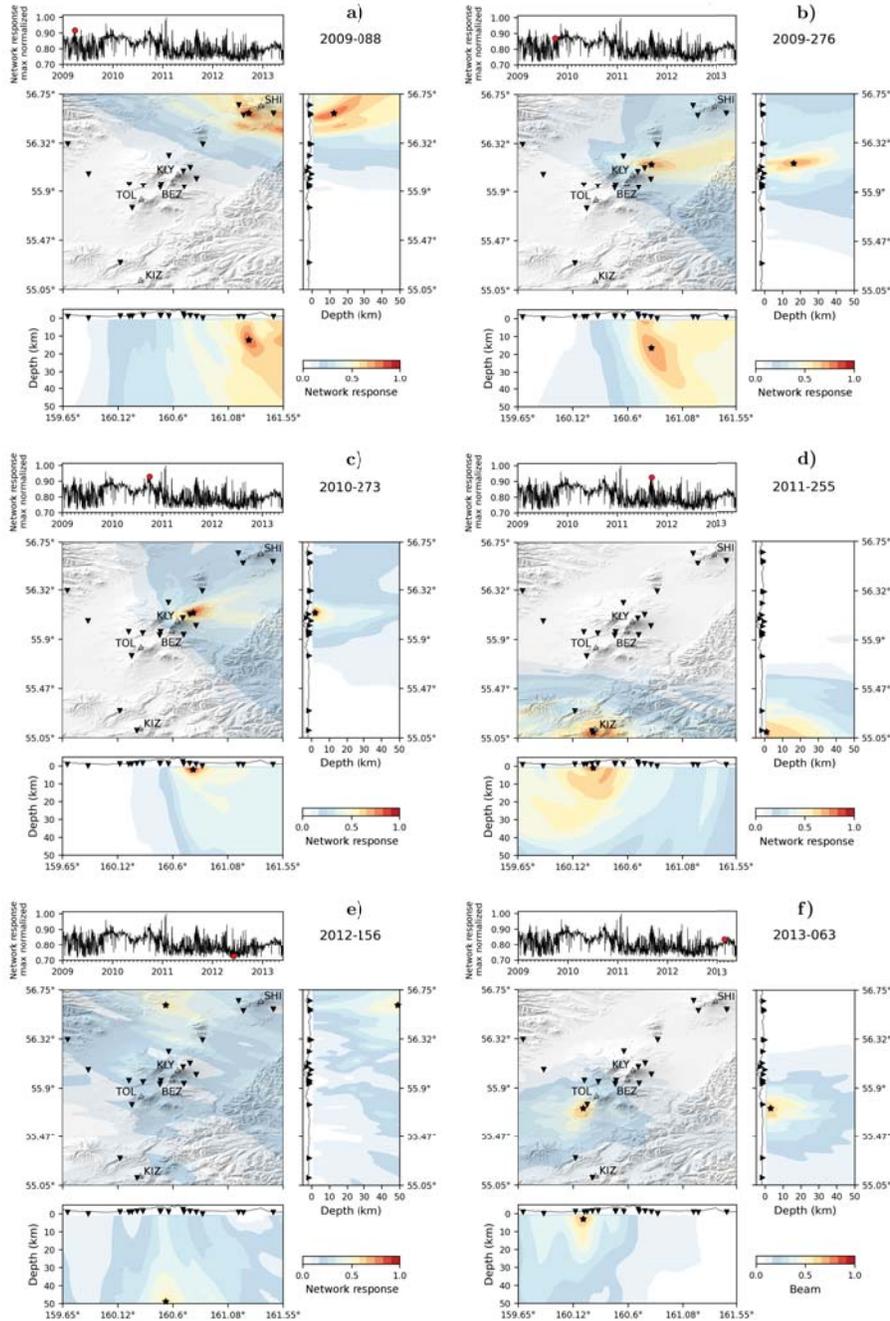
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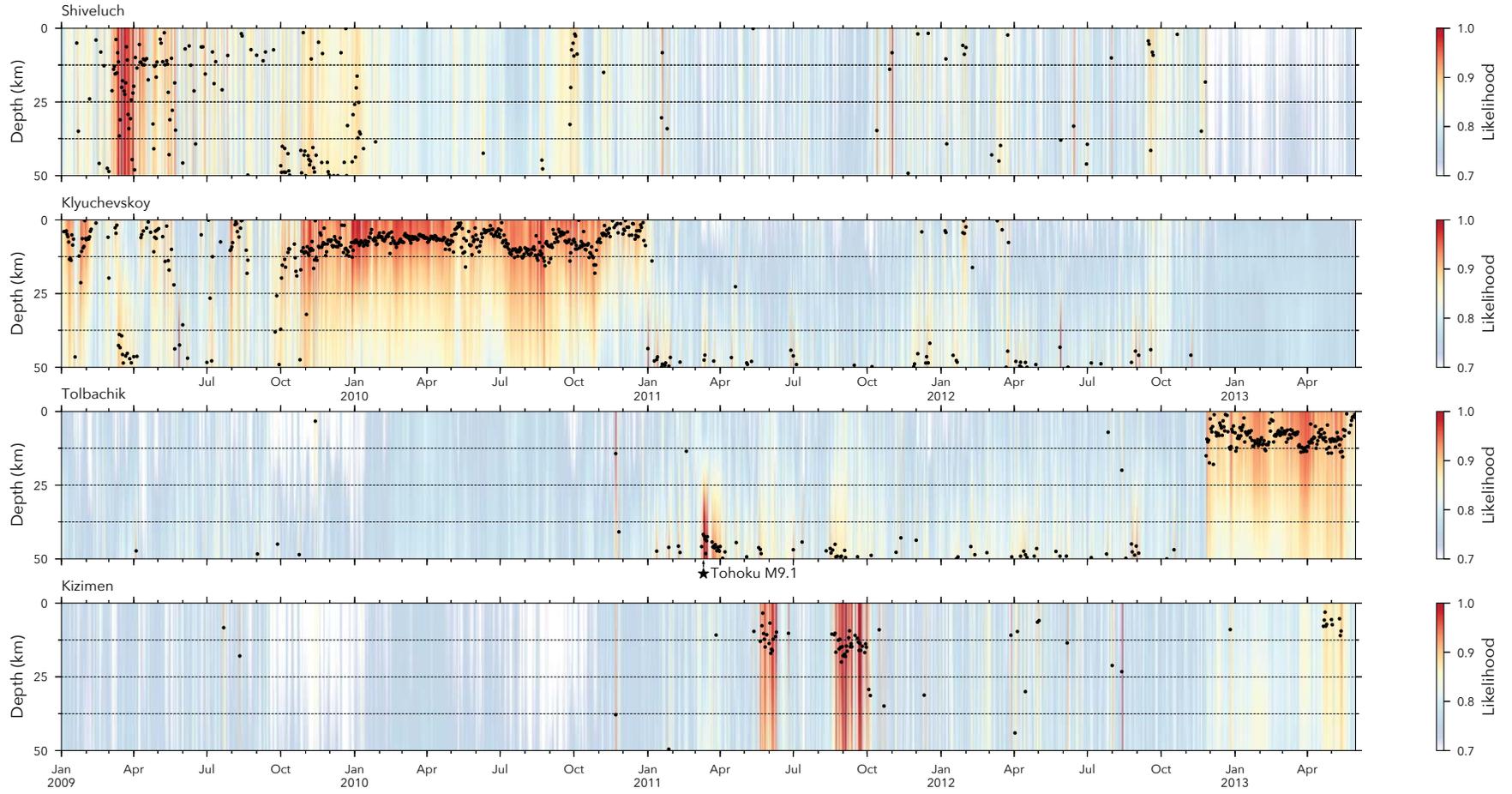
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**Figure 1.** Seismic network (stations: black inverted triangles) monitoring the Klyuchevskoy volcanic group and surrounding volcanoes (white triangles). The Russian Kamchatka peninsula is located by a red square on the Earth globe. The 25 km radius circular zones used for the analysis of the vertical migration of volcanic tremor sources with time addressed in Sec. 3.3 appear as red zones on the map. Note that those circular red zones appear as ellipses here because of the Miller projection (a modified Mercator projection) used for drawing the map.



**Figure 2.** Volcanic tremor source presence likelihood (corresponding to (11)) corresponding to six different days (red dot in top-left black curves). On each response function, the most probable tremor source location is marked by a black star. Each response function appears as a top-view of its values in the horizontal plane corresponding to the depth of the tremor source (center panel), and two side-views of its values in vertical planes corresponding respectively to the depth-longitude plane (bottom panel) and the depth-latitude plane (right panel) of the tremor source. Seismic stations and volcanoes respectively appear as black inverted triangles and with triangles in the horizontal top-view (center panel), and first three initial letters of each volcano help the reader to orient. **a)** Quite deep tremor beneath Shiveluch on day 2009-088. **b)** Deep tremor beneath Klyuchevskoy on day 2009-276. **c)** Shallow tremor beneath Klyuchevskoy on day 2010-273.



**Figure 3.** Time evolution of tremor sources depth beneath each studied volcano (Shiveluch, Klyuchevskoy, Tolbachik and Kizimen from North to South). A circular zone of radius  $r = 25$  km is defined around each volcano (the zone is centered on volcanoes summit as appearing on Fig. 1) and the maximum value of the likelihood (11) over those circular zones is selected at every depth and every day. The expected value  $\bar{z}$  of the likelihood for each zone and each time is calculated with  $\bar{z}(t) = \int \max_{x,y} \tilde{\ell}(\sqrt{x^2 + y^2} < r, z, t) z dz$  and represented as a black dot in order to better visualize the expected position of the tremor sources. The time of the 2011 M9.1 Tohoku earthquake (Ozawa et al., 2011) is indicated with a black star beneath the results from the Tolbachik volcano.

